LEE-SIDE FRONTOGENESIS IN THE ROCKY MOUNTAINS [1]

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ABSTRACT

The development and motion of fronts associated with lee-side troughs on large mountain barriers has been investigated. These fronts differ from ordinary cold fronts in their horizontal temperature field, which is characterized by a sinusoidal thermal ridge. The thermal ridge intensifies, while remaining stationary with respect to the mountains, and moves eastward upon the approach of a Pacific cold front.

An equation is derived, showing that changes in the thermal pattern can be described by changes in a potential thermal vorticity equation, which consists of three terms: (1) one representing an advection of the potential thermal vorticity by the 500-mb. wind; (2) one representing the advection of 500-mb. absolute vorticity by the thermal wind; and (3) a purely orographic term.

An idealized sinusoidal model of the thickness pattern is used in conjunction with the prognostic equation to explain the development and motion of lee-side thermal ridges. Actual examples from synoptic maps are chosen to corroborate the theory. The conclusions are: (1) the thermal ridge will develop when the surface flow is such as to produce large-scale descent on the lee slopes of the mountains; (2) no thermal ridging will appear when the 500-mb. ridge lies east of the lee slopes; (3) thermal ridging will appear with the approach of a 500-mb. ridge from the west; and (4) the thermal ridge will move eastward upon the passage of the 500-mb. ridge.

1. INTRODUCTION

A well-known phenomenon is the lee-side trough, associated with large mountain barriers. At certain times of the year low-pressure troughs frequently develop on the eastern slopes of the Rocky Mountains. A particular type of front can be found along the axis of the trough, which is distinguishable from ordinary cold or warm fronts by its horizontal temperature field. This kind of front will be called a "pseudo front" in this paper. Here the word "pseudo" is not to be confused with other definitions involving the term, as applied to squall lines and other meso-features [1]. Typically, cold fronts lie along the axes of low-pressure troughs, moving from colder to warmer air. In a typical cold front the thermal gradient is quite pronounced behind the front (fig. 1). Pseudo fronts differ from cold fronts in that (1) the associated surface trough does not show an abrupt temperature discontinuity at low levels, although the wind shift across the front may be quite pronounced, and (2) the thermal contours are almost symmetrical about the surface trough axis, with a maximum temperature at the front, decreasing at about the same rate on either side of the trough. It is evident that this description is exactly that of the temperature field associated with an occluded front. Pseudo fronts are distinguished from occlusions on the

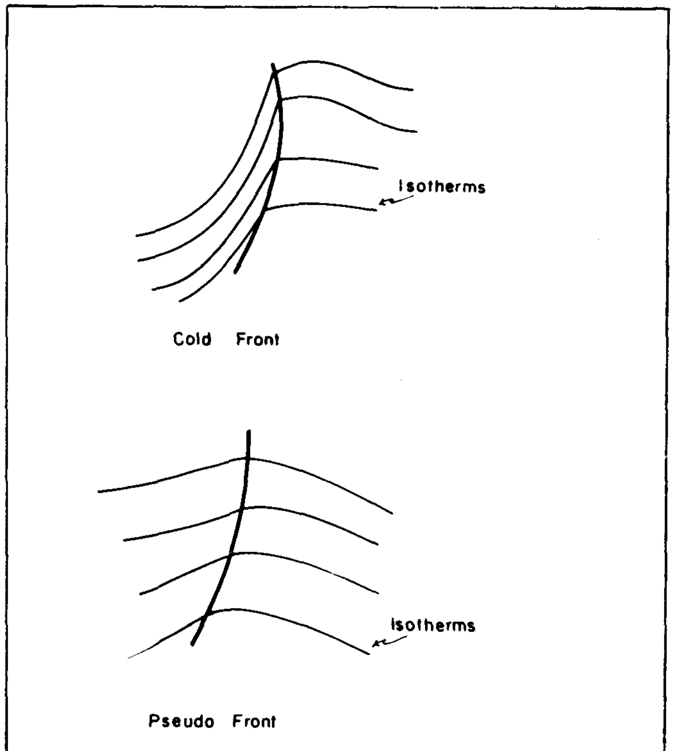

FIGURE 1.—Schematic diagrams of cold and pseudo fronts.

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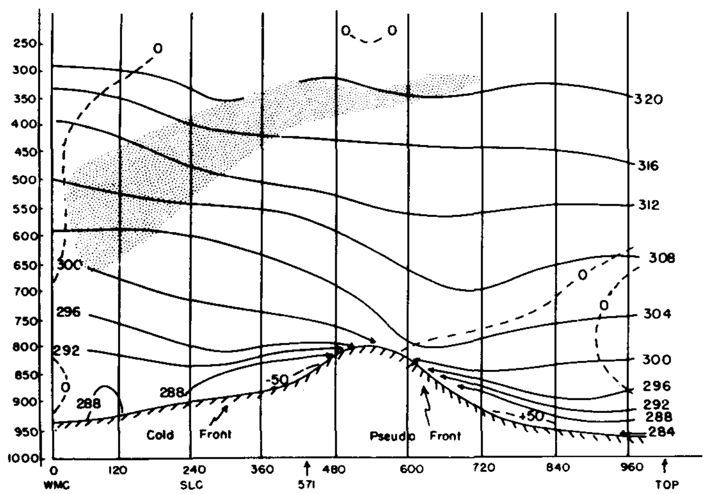

FIGURE 2.—Cross-section, Winnemucca, Nev. (WMC), to Topeka, Kans. (TOP), 1200 GMT, January 13, 1959. Solid lines are isotherms of potential temperature; dashed lines, vertical velocity, mb. (12 hr.)$^{-1}$. Shading indicates greater than three-eighths cloud cover. Distance is in miles.

basis of the development of the latter in the life of a frontal wave. In the early stages of cyclogenesis, the thermal structure of a pseudo front often resembles that of an occlusion, except that the system is intensifying. There seems to be no evidence in the historical continuity of pseudo fronts that a cold front has overtaken a warm front, such as occurs toward the end of the evolutionary cycle of a classical Bergeron cyclone.

The most common type of pseudo front on the North American continent develops along the lee slopes of the Rocky Mountains, particularly during the months of November through March. The front first appears in the thermal pattern as a sinusoidal ridge, with a wavelength of about 1500 km. At the surface, a marked wind shift line appears coincident with the thermal ridge axis. The wave intensifies in both the thermal and surface circulations, often becoming extended along the entire eastern slopes of the mountains, although during initial development the forward advance is often very slight. McClain [2] and Hess and Wagner [3] have studied in detail the development of lee-side cyclogenesis, and their analyses of 12-hour synoptic maps showed that the lee cyclones began to move eastward upon the approach of a trough from the Pacific. Cloudiness and precipitation accompanying the front are noticeably absent [4]. A cross-section through both the approaching cold front from the Pacific and the pseudo front at 1200 GMT, January 13, 1959, reveals not only the total lack of middle cloudiness and precipitation accompanying the pseudo front, but points out its distinct identity in the temperature field (fig. 2). The phenomenon of the lee-side pressure trough is of such frequent occurrence in the Rocky Mountain region, that mean sea level charts show it clearly [3]. The time required for full development of a lee-side trough is about 36 hours, and the rapid intensification of the thermal ridge east of the Rocky Mountains is associated with the commonly experienced "chinook" or "foehn." The ultimate eastward movement of the pseudo front seems to be influenced by approaching Pacific troughs.

The purpose of this paper is to try to explain the development and movement of these pseudo fronts in terms of the large-scale physical processes in the atmosphere. Similar work has been done by McClain [2], in which he correlated upper tropospheric divergence with the development of the lee-side troughs. Other researchers such as Colson [5] and Newton [6] have tried to explain the frequency of cyclogenesis in the central United States in terms of air flow over the Rocky Mountains.

2. THEORY

Since pseudo fronts are best revealed in the low-level temperature pattern, it becomes necessary to discuss the motion of the accompanying thermal ridge. It would be best to derive a comprehensive prediction equation for the thickness pattern, from the basic laws governing the atmosphere. One approach, which works reasonably well and is simple, is use of a two-parameter model which describes the atmosphere in terms of the mean wind and a thermal wind, taken over an appropriate depth in the atmosphere. The following modeling assumption is chosen:

$$\mathbf{V}(x, y, p, t)=\mathbf{V}_m(x, y, t)+B(p)\mathbf{V}_{m0}(x, y, t) \tag{1}$$

where $\mathbf{V}$ refers to the non-divergent component of the wind, and B is a function of pressure only. The definition of the mean quantities is written

$$(\ )_m\equiv\frac{1}{p_0}\int_0^{p_0}(\ )dp. \tag{2a}$$

The subscripts m and 0 respectively refer to the mean and 1000-mb. levels, where

$$(\ )_{m0}\equiv(\ )_m-(\ )_0. \tag{2b}$$

From the previous definitions, it is evident that $B_m=0$ and that the vertical shear is constant in a particular air column. Applying the curl operator to (1) gives the corollary

$$\zeta=\zeta_m+B\zeta_{m0}, \tag{3}$$

where ζ is the relative vorticity component in the vertical. Introduction of the simplified vorticity equation

$$\frac{\partial\zeta}{\partial t}=-\mathbf{V}\cdot\nabla(\zeta+f)+f\frac{\partial\omega}{\partial p}, \tag{4}$$

where ∇ is the gradient operator at constant pressure, $\omega\equiv\frac{dp}{dt}$, and f is the Coriolis parameter, yields an expression for large-scale atmospheric motion. The simplification refers to the deletion of the twisting, vertical advection, and frictional terms, and the assumption that $f \gg \zeta$. The boundary conditions for the model atmosphere are:

$$\omega(p=0)=0, \tag{5a}$$

$$\omega(p=p_s)=\omega_s \tag{5b}$$

where subscript s refers to the ground surface. Substitution of (1) and (3) in (4) yields

$$\frac{\partial \zeta_m}{\partial t}+B\frac{\partial \zeta_{m0}}{\partial t}=-\mathbf{V}_m\cdot\nabla(\zeta_m+f)-B\mathbf{V}_m\cdot\nabla\zeta_{m0}$$
$$-B\mathbf{V}_{m0}\cdot\nabla(\zeta_m+f)-B^2\mathbf{V}_{m0}\cdot\nabla\zeta_{m0}+f\frac{\partial\omega}{\partial p}. \tag{6}$$

To eliminate $\frac{\partial \zeta_m}{\partial t}$, apply (2) to (4), obtaining another expression for $\frac{\partial \zeta_m}{\partial t}$ and subtract the result from (6). Thus,

$$\frac{\partial \zeta_{m0}}{\partial t}=-\mathbf{V}_{m0}\cdot\nabla(\zeta_m+f)-\left\{\mathbf{V}_m+\mathbf{V}_{m0}\left[\frac{B^2-(B^2)_m}{B}\right]\right\}\cdot\nabla\zeta_{m0}$$
$$+\frac{f}{B}\left(\frac{\partial\omega}{\partial p}\right)-\frac{f}{B}\frac{\omega_s}{p_s}. \tag{7}$$

Application of this equation at the level where $B^2=(B^2)_m$ and defining $B_*\equiv\sqrt{(B^2)_m}$ yields:

$$\frac{\partial \zeta_{m0}}{\partial t}=-\mathbf{V}_{m0}\cdot\nabla(\zeta_m+f)-\mathbf{V}_m\cdot\nabla\zeta_{m0}+\frac{f}{B_*}\left(\frac{\partial\omega}{\partial p}\right)_*-\frac{f}{B_*}\frac{\omega_s}{p_s}. \tag{8}$$

Terms containing ω can be eliminated by introduction of the thermodynamic considerations. For a temperature field where the potential temperature $\theta=\theta(x,y,p,t)$,

$$\frac{d\theta}{dt}=\frac{\partial\theta}{\partial t}+\mathbf{V}\cdot\nabla\theta+\omega\frac{\partial\theta}{\partial p}. \tag{9}$$

The quantity θ is defined in the usual manner where

$$\theta\equiv T\left(\frac{p_0}{p}\right)^{\kappa}=\frac{\alpha}{R}\frac{p_0^{\kappa}}{p^{\kappa-1}}, \tag{10}$$

where T is the absolute temperature, α the specific volume, R the gas constant for dry air, $p_0=1000$ mb., and $\kappa\equiv R/c_p$. Since $\left(\frac{\partial\theta}{\partial t}\right)_p=\frac{\theta}{\alpha}\left(\frac{\partial\alpha}{\partial t}\right)$ and $\nabla\theta=\theta/\alpha\nabla\alpha$, equation (9) becomes, after multiplication by α/θ and substitution of these relationships

$$\frac{\alpha}{\theta}\frac{d\theta}{dt}=\frac{\partial\alpha}{\partial t}+\mathbf{V}\cdot\nabla\alpha+\frac{\alpha}{\theta}\frac{\partial\theta}{\partial p}\omega. \tag{11}$$

The requirement that the atmosphere be in hydrostatic equilibrium gives

$$\alpha=-g\frac{\partial z}{\partial p}, \tag{12}$$

where g is the gravitational constant and z is the height of a constant pressure surface. Substitution of (12) into (11) yields, upon division by g and rearrangement of terms

$$\frac{\partial}{\partial t}\left(\frac{\partial z}{\partial p}\right)=-\mathbf{V}\cdot\nabla\left(\frac{\partial z}{\partial p}\right)+\frac{\alpha}{g\theta}\frac{\partial\theta}{\partial p}\omega-\frac{\alpha}{g\theta}\frac{d\theta}{dt}. \tag{13}$$

Assuming that no non-adiabatic heating occurs, integration of (13) with respect to pressure from 1000 mb. to the mean level, where the pressure is p_m, yields

$$\frac{\partial z_{m0}}{\partial t}=-\mathbf{V}_m\cdot\nabla z_{m0}+\int_{1000}^{p_m}\frac{\alpha}{g\theta}\frac{\partial\theta}{\partial p}\omega dp, \tag{14}$$

since $\mathbf{V}_{m0}\cdot\nabla\left(\frac{\partial z}{\partial p}\right)=0$ and $\mathbf{V}\cdot\nabla\left(\frac{\partial z}{\partial p}\right)=\mathbf{V}_m\cdot\nabla\left(\frac{\partial z}{\partial p}\right)$. From the equation of continuity

$$\omega=-\int_0^p\nabla\cdot(\mathbf{V}_m+B\mathbf{V}_{m0})\,dp=\frac{p}{p_s}\omega_s-\nabla\cdot\mathbf{V}_{m0}\int_0^p Bdp. \tag{16}$$

Similarly

$$\left(\frac{\partial\omega}{\partial p}\right)_*=\frac{\omega_s}{p_s}-B_*\nabla\cdot\mathbf{V}_{m0}. \tag{17}$$

By substitution of (16) in (14), and defining $c\equiv\int_0^p Bdp$ the thermodynamic equation becomes

$$\frac{\partial z_{m0}}{\partial t}=-\mathbf{V}_m\cdot\nabla z_{m0}+\frac{\omega_s}{p_s}E-\nabla\cdot\mathbf{V}_{m0}D, \tag{18}$$

where $E\equiv\int_{1000}^{p_m}\frac{\alpha p}{g\theta}\frac{\partial\theta}{\partial p}dp$, and $D\equiv\int_{1000}^{p_m}\frac{c\alpha}{g\theta}\frac{\partial\theta}{\partial p}dp$. Substitution of (17) into (8) yields

$$\frac{\partial\zeta_{m0}}{\partial t}=-\mathbf{V}_{m0}\cdot\nabla(\zeta_m+f)-\mathbf{V}_m\cdot\nabla\zeta_{m0}-f\nabla\cdot\mathbf{V}_{m0}. \tag{19}$$

Elimination of $\nabla\cdot\mathbf{V}_{m0}$ between (18) and (19), and rearrangement of terms yields

$$\frac{\partial}{\partial t}\left(\zeta_{m0}-\frac{f}{D}z_{m0}\right)=-\mathbf{V}_m\cdot\nabla\left(\zeta_{m0}-\frac{f}{D}z_{m0}\right)$$
$$-\mathbf{V}_{m0}\cdot\nabla(\zeta_m+f)-\frac{fE\omega_s}{p_sD}, \tag{20}$$

which can be written in the abbreviated form

$$\frac{\partial}{\partial t}(\zeta_{m0}-Kz_{m0})=-\mathbf{V}_m\cdot\nabla(\zeta_{m0}-Kz_{m0})-\mathbf{V}_{m0}\cdot\nabla(\zeta_m+f)-k\omega_s, \tag{21}$$

where $K\equiv f/D$ and $k\equiv fE/(p_sD)$. E and D can be evaluated from Buch's [7] mean data, and from mean values of $\partial\theta/\partial p$ for western North America.

In a typical synoptic situation, changes in ζ_{m0} and $-Kz_{m0}$ augment each other since increases or decreases in thermal vorticity are associated, respectively, with decreases or increases in thickness. The combined expression $(\zeta_{m0}-Kz_{m0})$ can be called a potential thermal vorticity. This means that the thermal vorticity ζ_{m0} has a potential value with respect to some value of z_{m0}. Using the symbol ϕ for the potential thermal vorticity, expression (21) becomes

$$\frac{\partial\phi}{\partial t}=-\mathbf{V}_m\cdot\nabla\phi-\mathbf{V}_{m0}\cdot\nabla(\zeta_m+f)-k\omega_s, \tag{22}$$

or

$$\frac{d\phi}{dt}=-\mathbf{V}_{m0}\cdot\nabla(\zeta_m+f)-k\omega_s. \tag{23}$$

The terms $-\mathbf{V}_{m0}\cdot\nabla(\zeta_m+f)$ and $-k\omega_s$ represent nonconservative, or generative, effects. The presence of generative terms suggests a means by which thermal ridges are suddenly built up on the lee slopes of large mountain ranges. The former term is not localized geographically, so that $-k\omega_s$ must represent the effect which is responsible for the lee-slope development. It is conceivable that the orographic contribution can influence the thermal pattern in such a way that $-\mathbf{V}_{m0}\cdot\nabla(\zeta_m+f)$ itself is altered, making that a secondary effect of the orography. Since ω_s is positive for winds blowing from higher to lower levels, downslope winds will always contribute negatively to $\partial\phi/\partial t$ and increase thermal ridging.

The entire expression (22) is quite consistent with other types of manipulation of the two-parameter model atmosphere, such as those employed by Sawyer and Bushby [8] and Sutcliffe [9], but is more appropriate for the present problem in that it refers explicitly to the thermal pattern. Equation (21) is not limited to orographic effects, and can be utilized to study any kind of large-scale thermal pattern. Consider a typical mid-latitude trough-ridge pattern at the mean level, where a low-level baroclinic cyclone appears on the eastern side of the upper trough. The isotherms associated with the surface Low will advance northward, becoming greatly extended, and finally become concentrically closed. Because of the close association between changes in the potential thermal vorticity and changes in the thickness, the first term on the right side of (21) explains the northward movement of the thermal pattern. Now, the origin of closed isotherms cannot be explained by a single horizontal advection process, and therefore, a second effect, such as that described by the last term on the right side of (23), must be responsible for the development of pools of cold and warm air. For this reason equation (21) could be appropriate in studying cold Lows and warm Highs.

3. THE IDEALIZED MODEL

Consider an ideal sinusoidal z_{m0} field as schematically represented in figure 3. The x-axis lies in the east-west direction, and the y-axis coincides with the north-south mountain ridge. The crest of the idealized thermal ridge coincides with x''. The following statement can then be made

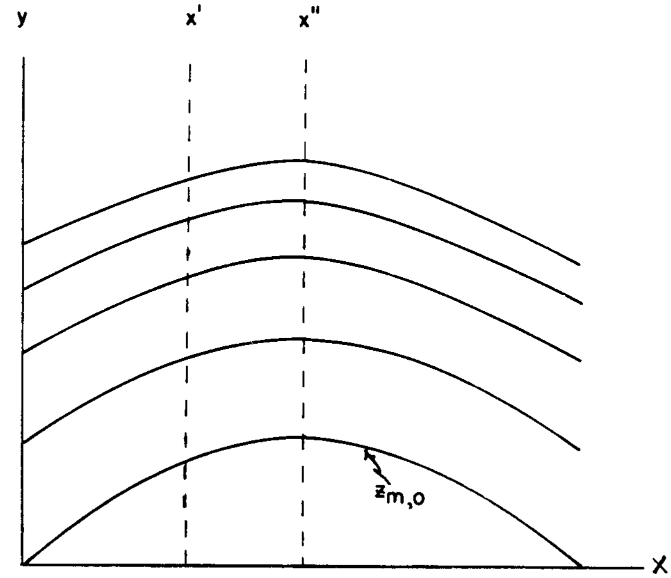

FIGURE 3.—Schematic model of ideal thickness pattern. Isolines of z_{m0} in the x, y plane. x''=axis of ridge. x'= position of maximum descent.

$$z_{m0}=a\sin\frac{2\pi x}{\lambda}+by, \tag{24}$$

where a is the amplitude of the ridge, λ the wavelength of the pattern, and b is taken to be a constant, although it generally slightly increases toward the north. Now if the geostrophic assumption holds,

$$\zeta_{m0}=\frac{g}{f}\nabla^2 z=-\frac{g}{f}\left(\frac{2\pi}{\lambda}\right)^2 a\sin\frac{2\pi x}{\lambda}. \tag{25}$$

The above expression indicates the resemblance between ϕ and z_{m0} since

$$\phi=a'\sin\frac{2\pi x}{\lambda}+b'y, \tag{26a}$$

where

$$a'\equiv-\left[\frac{g}{f}\left(\frac{2\pi}{\lambda}\right)^2+K\right]a,\text{ and } b'\equiv-K\,b. \tag{26b}$$

Consider a pattern of flow at the mean level, where the component of the mean wind $u_m\equiv U$, and v_m is represented by a perturbation v', and the term, $-\mathbf{V}_{m0}\cdot\nabla(\zeta_m+f)$ is represented by the symbol Q. The 500-mb. pattern consists of a steady zonal current with a speed U upon which is superimposed a transversal wave perturbation. The expression (21) becomes

$$\frac{\partial \phi}{\partial t}=-U\frac{\partial \phi}{\partial x}-v'\frac{\partial \phi}{\partial y}-k\omega_s+Q. \tag{27}$$

In order to determine the eastward movement of the thermal ridge, this equation is differentiated with respect to x at the point x'':

$$\frac{\partial}{\partial x}\left(\frac{\partial \phi}{\partial t}\right)_{x''}=-U\left[\left(\frac{2\pi}{\lambda}\right)^2\left\{\frac{g}{f}\left(\frac{2\pi}{\lambda}\right)^2+K\right\}\right]a -k\frac{\partial \omega_s}{\partial x}+\frac{\partial v'}{\partial x}K\,b+\frac{\partial Q}{\partial x}. \tag{28}$$

The last two terms on the right side of (28) represent the influence on the motion of the thermal ridge by the perturbed part of the mean flow, and therefore demonstrate the effects of migratory troughs and ridges at the mean level. It is evident that $\partial v'/\partial x$ is merely ζ_m, and that the entire term, $Kb\zeta_m$, represents an advective influence by the non-zonal component of the mean wind. The last term represents a generating effect on thermal ridges by vorticity advection gradients at the mean level. Henceforth, these two terms will be combined as one, where $Q'\equiv Kb\zeta_m+\frac{\partial Q}{\partial x}$. From (28) the following conditions exist:

$$\frac{\partial}{\partial x}\left(\frac{\partial \phi}{\partial t}\right)_{x''}=0,\ \text{stationarity} \tag{29a}$$

$$\frac{\partial}{\partial x}\left(\frac{\partial \phi}{\partial t}\right)_{x''}>0,\ \text{quasi stationarity} \tag{29b}$$

$$\frac{\partial}{\partial x}\left(\frac{\partial \phi}{\partial t}\right)_{x''}<0,\ \text{eastward movement} \tag{29c}$$

First consider the equation of stationarity: solving for a at x'', (28) becomes

$$a=\frac{-k\left(\frac{\partial \omega_s}{\partial x}\right)+Q'}{U\left[\left(\frac{2\pi}{\lambda}\right)^2\left\{\frac{g}{f}\left(\frac{2\pi}{\lambda}\right)^2+K\right\}\right]}=a_c. \tag{30}$$

If the magnitude of Q' is less than the magnitude of $k\frac{\partial \omega_s}{\partial x}$, $\frac{\partial \omega_s}{\partial x}$ must be a negative quantity in order for a to be positive. Thus stationary thermal ridges must become positioned where $\frac{\partial \omega_s}{\partial x}$ is negative, in particular at a point downstream from the region of maximum descent. Equation (30) states that there exists a positive finite amplitude which will satisfy the stationarity condition. This amplitude will be called the "critical amplitude" represented by the symbol a_c. Substitution of (30) in (28) yields

$$\frac{\partial}{\partial x}\left(\frac{\partial \phi}{\partial t}\right)_{x''}=U\left[\left(\frac{2\pi}{\lambda}\right)^2\left\{\frac{g}{f}\left(\frac{2\pi}{\lambda}\right)^2+K\right\}\right](a_c-a). \tag{31}$$

Equation (31) explicitly states that the thermal ridge will advance eastward, if $a_c>a$.

Consider the idealized development of a thermal ridge when a steady pattern of ω_s is imposed upon the lee slope of the mountain at $t=0$. The initial position of the ridge will be at x', since the generating effect of $-k\omega_s$ is greatest at that point. Now $\partial\omega_s/\partial x$ is zero at x', and therefore, according to (30) and (31), the ridge must move eastward in the first instant, but as it does so, a_c also increases according to (30). The term, $-k\omega_s$, represents an effect which is continuously generating a thermal ridge, and thus increasing the amplitude of the ridge with time. The critical amplitude can never exceed the actual amplitude without imposing the stationarity condition. Similarly, the actual amplitude can never greatly exceed the critical amplitude, since that quantity will also increase as the thermal ridge moves east of x'. If $a\approx a_c$, equation (31) states that an approximate condition of stationarity exists. Although the extent and duration of approximate stationarity cannot be shown, the condition will persist until the thermal ridge attains a position downstream from x', where $\partial\omega_s/\partial x$ is at a minimum. Further eastward movement will be accompanied by a rapid decrease in critical amplitude according to (30), and a negation of the condition of approximate stationarity.

The horizontal advection of ϕ represents a decrease in this quantity ahead of x'', and an increase behind it, while $-k\omega_s$ is decreasing ϕ ahead of x'', but is decreasing it more strongly at x'. The growing importance of the advection of ϕ causes the ridge to move slowly eastward. The ultimate reduction of the critical amplitude corresponds physically to the domination of the horizontal advection over the orographic process.

If Q' is negative across the ridge axis, the numerator in (30) will tend toward zero, leaving $a_c\ll a$. According to (31), this event will tend to remove an approximate stationarity condition which may be present. If no ridge is present, none may be allowed to form under such circumstances, since any amplitude which is created will be immediately advected downstream. Similarly, when Q' is positive, equation (30) states that the maximum critical amplitude will be much larger than in the situation where Q' is negative. In effect, this means that the approximate stationarity condition exists for a considerably longer time.

Since b is a negative constant in the present reference system, from the definition of Q', negative relative vorticity at the mean level contributes to stationarity. The generative effect of Q will also contribute to stationarity when $\partial Q/\partial x$ is positive, or when stronger cyclonic vorticity advection occurs east of the thermal ridge. Calculations of this term have shown its magnitude to be very small, and therefore Q' primarily consists of ζ_m. Thus Q' will tend to be strongly positive in the vicinity of a 500-mb. ridge, and begin to decrease sharply behind it.

In conclusion, it must be pointed out that (1) the initial

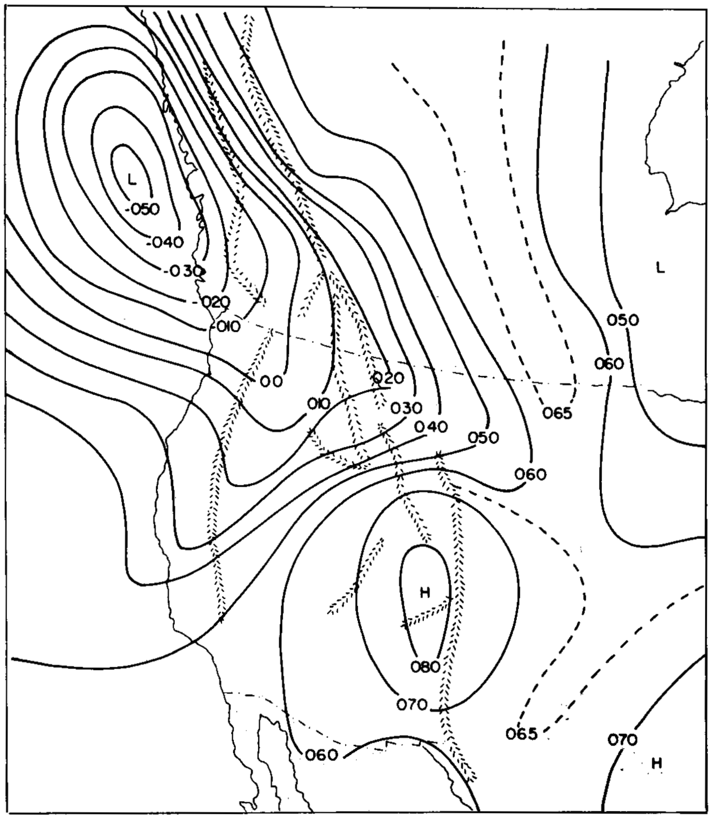

FIGURE 4.—1000-mb. height contours (ft.), 1200 GMT, January 12, 1959 (unit digits omitted).

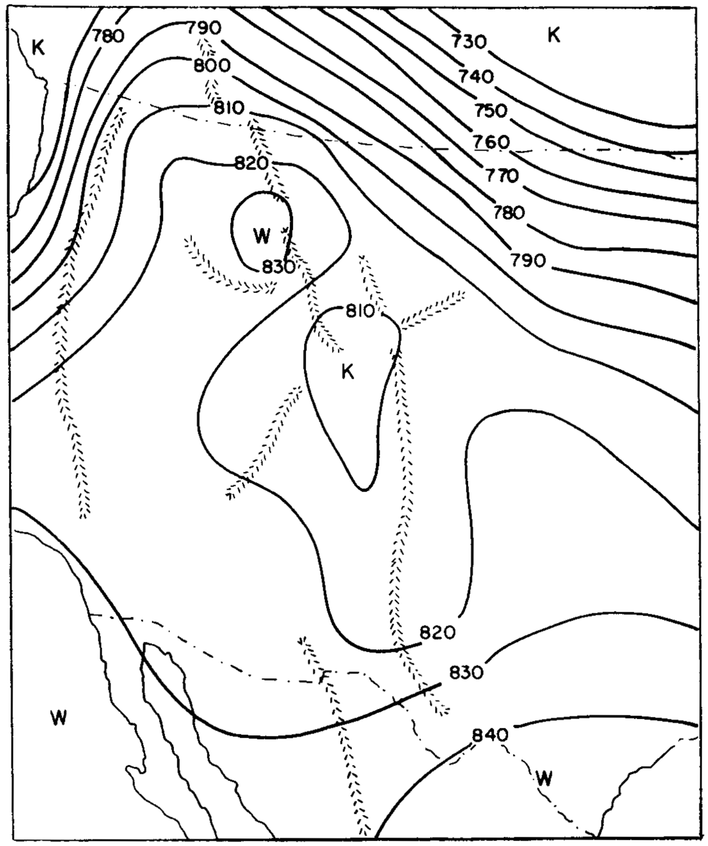

FIGURE 5.—1000–500-mb. thickness contours (ft.), 1200 GMT, January 12, 1959 (ten of thousands and unit digits omitted).

formation of a thermal ridge is dependent on the fundamental assumption that appreciable downslope velocities exist, and (2) the initial formation of the thermal ridge will not be observed as a lee-slope phenomenon if large positive values of relative vorticity at 500 mb. are occurring over the lee slopes of the mountain. If Q' is the largest in the numerator of (30) the position of the thermal ridge will not necessarily be localized on the east slope of the mountain range. Finally (3), the magnitude of both terms in the numerator of (30) will tend to vary with the wind speed, and therefore the critical amplitude may be somewhat insensitive to wind speed.

4. EXAMPLES OF LEE CYCLOGENESIS

A SPECIFIC CASE

An example of pseudo front development on the lee slopes of the Rocky Mountains has been chosen. The situation selected occurred between 1200 GMT, January 12, 1959, and 1200 GMT, January 13, 1959. The series of surface maps (figs. 4, 7, 10) reveals the development of the surface front. It was first noticed in connection with a feeble trough in the 065 contour in figure 4, and later exhibited a pronounced wind shift line, although no marked horizontal temperature contrast was present. The series of thickness maps (figs. 5, 8, 11) indicates the

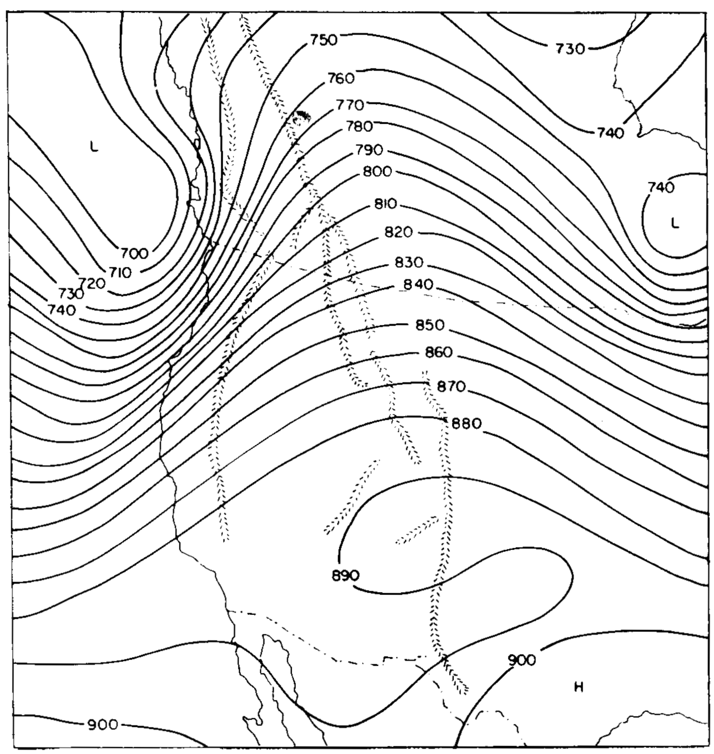

FIGURE 6.—500-mb. height contours (ft.), revealing strong cyclonic vorticity advection from the Pacific, 1200 GMT, January 12, 1959 (ten of thousands and unit digits omitted).

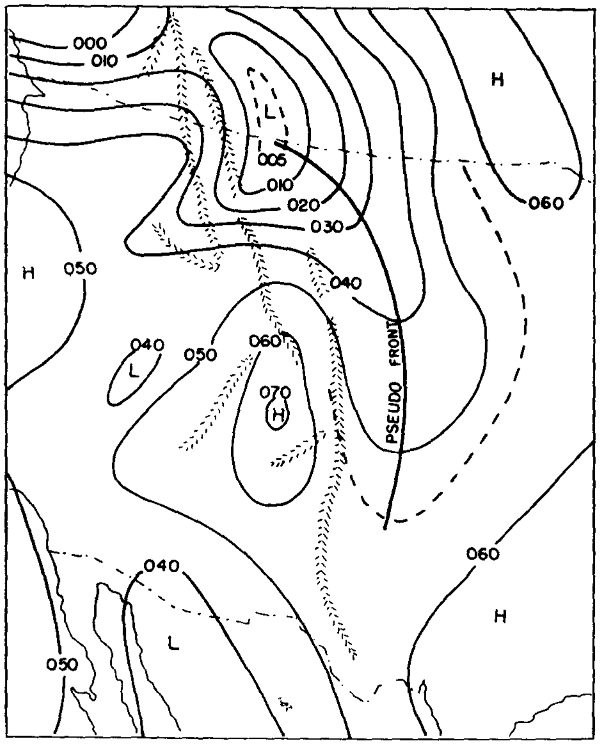

Figure 7.—1000-mb. height contours (ft.), 0000 GMT, January 13, 1959 (unit digits omitted).

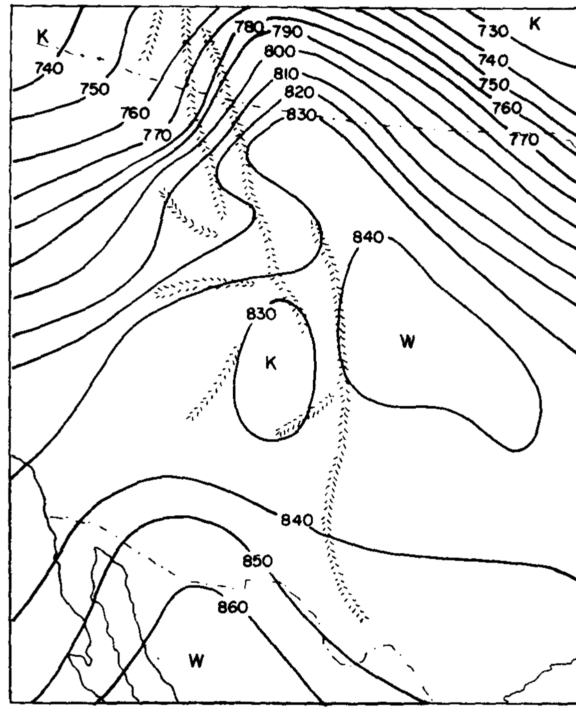

Figure 8.—1000–500-mb. thickness contours (ft.), 0000 GMT, January 13, 1959 (ten of thousands and unit digits omitted).

typical sinusoidal shape of the thermal pattern. Of note is the deep pocket of warm air around Colorado, associated with strong descending motion in that area. Although not presented here, maps preceding 1200 GMT, January 12, indicate the presence of a weak thermal ridge. Over Canada, the temperature field associated with the vigorous Pacific Low has apparently merged with the northern portion of the thermal ridge, and continued to move rapidly eastward. In terms of equation (30), the large cyclonic vorticity advection at 500 mb. (fig. 6) over southern Canada was removing the stationarity condition over the eastern slope of the mountain. South of the Canadian border, the 500-mb. ridge was positioned slightly west of the thermal ridge (figs. 9, 12). As the 500-mb. ridge moved east of the thermal ridge, the latter began to move eastward. This event could be anticipated by the idealized model, since Q' would tend to decrease sharply upon the passage of a 500-mb. ridge. Large Pacific Ocean cyclones which approach Puget Sound create strong westerly flow across a considerable part of the mountain chain. According to the model, this event occurring simultaneously with the approach of a 500-mb. ridge from the west, is ideal for the establishment of a stationary thermal ridge on the lee slopes.

GENERAL CASE

A general approach was utilized to study lee cyclogenesis. All instances of thermal ridge intensification on the lee slopes of the Rocky Mountains were systematically

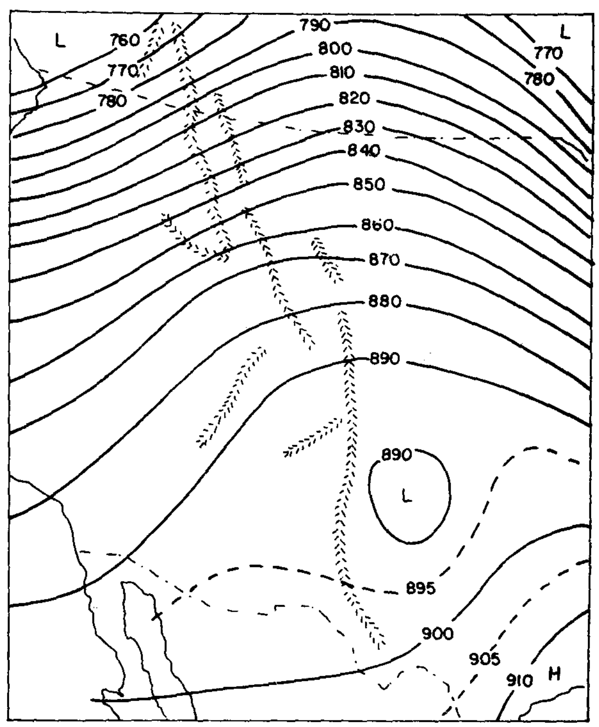

Figure 9.—500-mb. height contours (ft.), 0000 GMT, January 13, 1959 (ten of thousands and unit digits omitted).

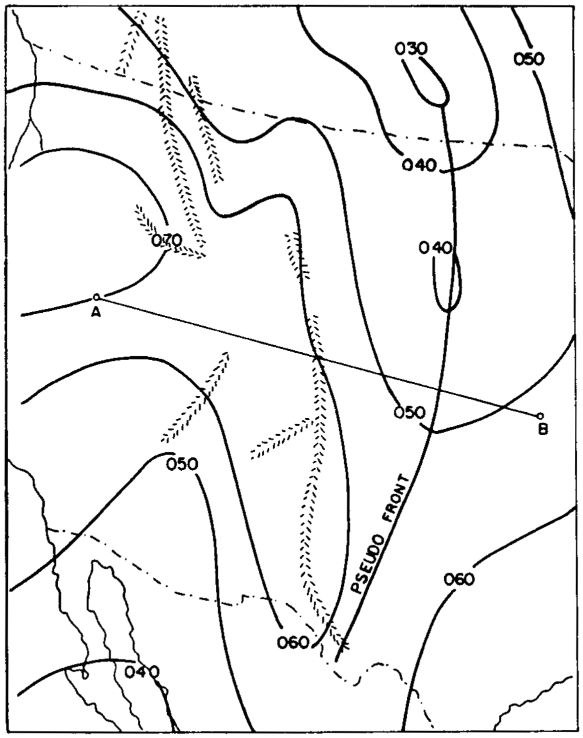

FIGURE 10.—1000-mb. height contours (ft.), 1200 GMT, January 13, 1959. A–B is position of atmospheric cross-section represented in figure 2 (unit digits omitted).

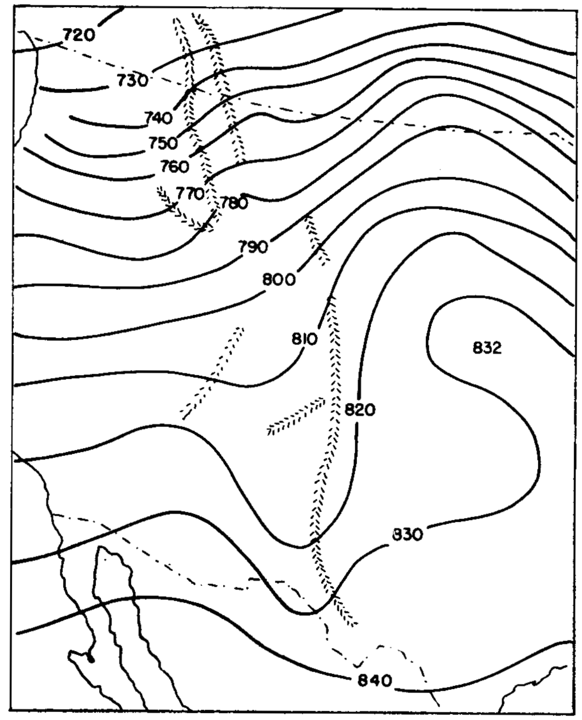

FIGURE 11.—1000–500-mb. thickness contours (ft.), 1200 GMT, January 13, 1959 (ten of thousands and unit digits omitted).

tabulated over the months December through March 1958–59 and 1959–60. The classification of the ridges was divided into two categories, "stationary" and "moving". The decision concerning the category of motion was qualitative, and based upon whether the thermal ridge appreciably moved eastward during the following 12-hour map interval.[2] Thermal ridges which remained stationary on more than one successive map were tabulated as "stationary" more than once. If the ridge later began to move eastward, it was tabulated again as "moving." For each case the position of the 500-mb. ridge and trough, relative to the thermal ridge, was tabulated. The results of this undertaking are presented in figure 13. The outstanding feature is the shift of the mean position of the 500-mb. ridge in the "stationary" cases to the east of the thermal ridge in the "moving" ones. Similarly, the mean position of the 500-mb. trough is situated at a distance greater than 15° of latitude west of the "stationary" thermal ridges, and about 12° west of the "moving" ones. The mean direction of the 500-mb. wind across the axis of the thermal ridge was 290° for the "stationary" cases and 265° in cases of "moving" thermal ridges. In all of the cases of thermal ridge development and motion (about 55), only four instances were accompanied by no definite trough-ridge pattern at 500 mb. The importance

[2] The surface map was always categorized before the 500-mb. flow was observed and categorized.

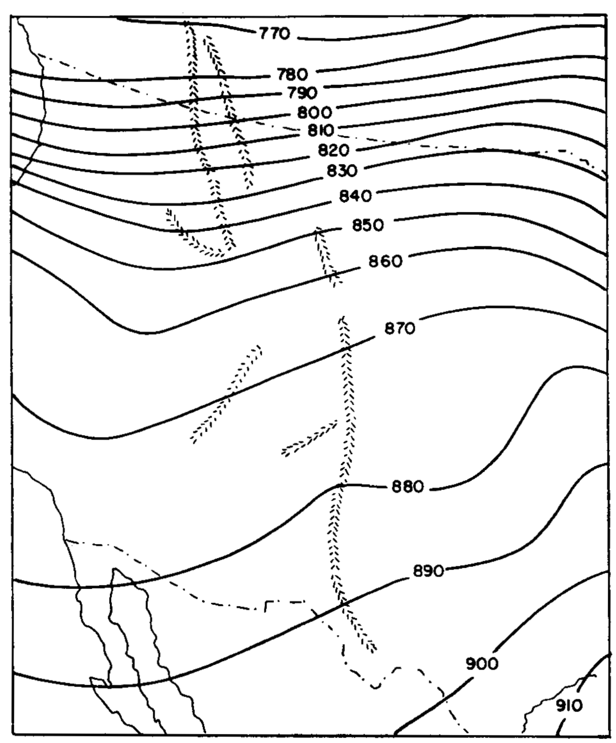

FIGURE 12.—500-mb. height contours (ft.), 1200 GMT, January 13, 1959 (ten of thousands and unit digits omitted).

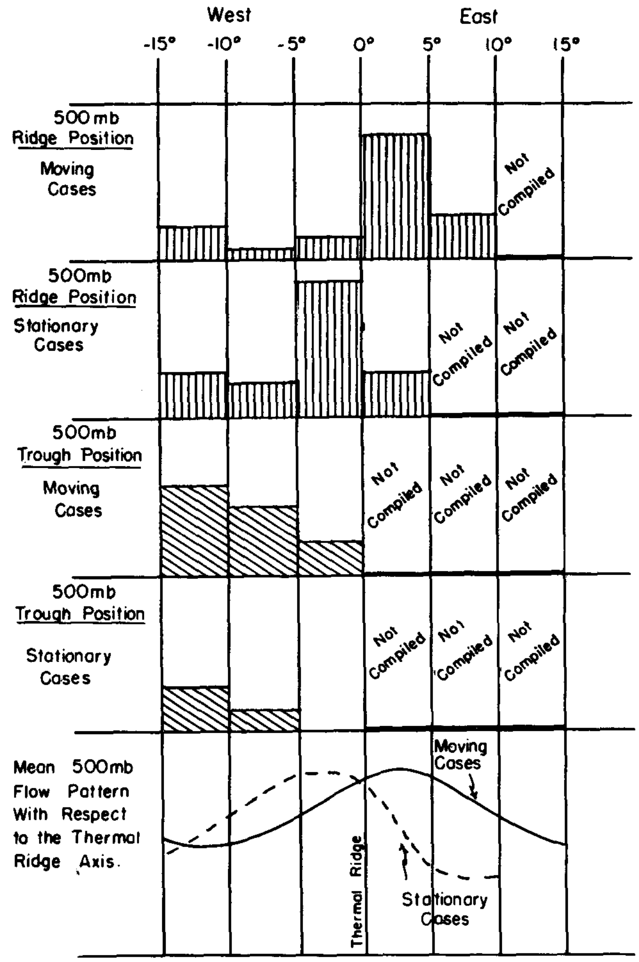

FIGURE 13.—Occurrence of 500-mb. troughs and ridges in 5° latitude intervals with respect to the thermal ridge axis, for stationary and moving cases. Compiled for the winters of 1958–59 and 1959–60 (units of ordinate go from 0 to 12 cases).

of the 500-mb. ridge in influencing the motion of lee-side disturbances is expressed by the term Q' in the numerator of equation (31). As the 500-mb. ridge approaches the lee slopes of the mountains, not only are conditions favorable for the initial development of thermal ridges, but those ridges which form will remain stationary for a short time due to the increasing effect of Q' which offsets the tendency for the thermal ridge to be advected downstream. Figure 13 also suggests that in the vicinity of 500-mb. troughs the numerator of (31) becomes vanishingly small, preventing the formation of thermal ridges.

5. SUMMARY AND CONCLUSIONS

The ultimate purpose of this paper was to study the development and motion of the frequently observed lower tropospheric frontogenesis associated with thermal ridge intensification on the lee slopes of the Rocky Mountains. The dynamical formulation showed that changes in the thermal pattern can be described by a potential thermal vorticity equation, which consists of three terms: (1) a term representing the advection of the potential thermal vorticity by the mean wind; (2) an orographic term; and (3) a term representing the advection of absolute vorticity at the mean level by the thermal wind. The mean level was considered to be 500 mb. An idealized sinusoidal model of the thermal ridge was devised in order to demonstrate the effect of each of the three terms on its motion. The results showed that the orographic effect was likely responsible for the thermal ridge amplification on the lee slopes of the mountains. According to the idealized pattern, the thermal ridge remains approximately stationary downstream from the point of maximum descent during the intensification process. Its amplitude continues to intensify until it attains a maximum critical value. At this point, the advection of potential thermal vorticity by the zonal flow at 500 mb. begins to dominate the motion, and the thermal ridge accelerates eastward. The effects of the perturbation in the 500-mb. current were shown to highly influence the development and motion of thermal ridges. The presence of migratory troughs and ridges at 500 mb. either (1) increases the maximum critical amplitude for stationarity, and thus increases the length of time for development, or (2) decreases the maximum critical amplitude, and therefore removes the stationarity condition, or (3) prevents any localized lee-slope thermal amplification from occurring.

An actual case was chosen to examine the development and motion of a thermal ridge in terms of the dynamical considerations. This undertaking provided a link between the theory and actual synoptic experience.

A survey of many cases was taken over the period of two winters. The results presented support the theory. Almost every case selected, whether it was stationary or moving eastward, was associated with a ridge at 500 mb. In the moving cases the 500-mb. ridge was generally centered east of the thermal ridge, and in the stationary instances it was centered west of the thermal ridge. Combined with the theory, these results lead to some conclusions concerning the nature of pseudo front development and motion.

The forecaster should consider the following criteria: (1) The thermal ridge will develop when the surface flow is such as to produce large-scale descent on the lee slopes of the mountains. (2) No thermal ridging will take place when the 500-mb. ridge lies downstream from lee slopes. (3) Thermal ridging will appear with the approach of a 500-mb. ridge from the west, and (4) will move eastward upon the passage of the 500-mb. ridge.

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